DISTRIBUTION OF POSITIVE IONS IN THE F-REGION

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ABSTRACT. The distribution of positive ions in the F-region of the ionosphere is controlled mainly by the following processes:

(*a*) photo-ionization of atmospheric constituents,
(*b*) ion-atom interchange,
(*c*) dissociative recombination of molecular ions with electrons and
(*d*) diffusion and drift of ionization.

Of these processes, the rate of production of ions from photoionization can be calculated with a fair degree of accuracy from rocket data (Hinteregger, 1961) of the solar spectrum in the ultraviolet and x-ray regions. The rate coefficients of dissociative recombination process are known but the rate coefficients of ion-atom interchange reactions, which are responsible for reshuffling the ions, are not available. From the rocket-borne experimental determination of percentage of ion compositions (Johnson, 1961), the rate coefficients of these reactions are obtained.

In order to reproduce the known ion composition, the rate coefficients of ion-atom interchange reactions should vary with altitude. Assuming that the temperature dependence of the above process is analogous to that of a two-body chemical reaction, their altitude variation and activation energies are obtained. The rate coefficient of the reaction,

$$O^+ + O_2 \rightarrow O_2^+ + O$$

at 300°K is found to be 3.07×10^{-11} cm³/sec which agrees with the experimental value of $2.5 \pm 0.4 \times 10^{-11}$ cm³/sec obtained by Dickinson and Sayers (1960) at the same temperature.

INTRODUCTION

To understand many upper atmospheric phenomena, e.g., night airglow, aurora, twilight airglow etc., the processes controlling the distribution of ions and electrons in the upper atmosphere should be clearly understood. It has been realised by many workers (Martyn, 1959, Yonezawa and Takahashi, 1960, Bates and Nicolet, 1960, Hertzberg, 1961) that the following processes together with the diffusion of ions and electrons are mainly controlling the distribution of ions in the ionosphere.

(1) photo-ionization of the atmospheric constituents,
(2) ion-atom interchange, and
(3) dissociative recombination of molecular ions with electrons.

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However, the relative contributions of these processes in controlling the distribution of ions in the ionosphere is not yet definitely known.

With the exception of the ion-atom interchange reactions, the rate coefficients of the above processes are known. In this paper from the rocket-borne experimental determination (Johnson, 1961) of the percentage of ion composition upto 220 km, the rate coefficients of these reactions are obtained. Using these rate coefficients, the ion distribution in the F-region is calculated upto 400 km. It is found that above 180 km, diffusion and ion drift must be taken into account to obtain the correct ion distribution with height.

DAYTIME IONOSPHERIC PROCESSES

The ionization of the upper atmospheric particles, which consists mainly of O_2, N_2, O, N and NO, is caused by solar radiations in the X-ray and ultraviolet regions. Rates of production of ions by X-ray have been calculated by Ghosh and Sharma (1961). In addition to ionization by X-rays, Ghosh and Shardanand (1960, 1961) considered ionization by ultraviolet radiations and that produced by Auger electrons and photoelectrons.

Bates (1955) showed that primary photo-ions are reshuffled by the ion-atom interchange process. To obtain daytime ion distribution in the F-region, consider the possible ion-atom interchange reactions as given by Hertzberg (1961), namely,

$$O^+ + N_2 \rightarrow NO^+ + N + 1.114 \text{ ev} \quad (1a)$$
$$O^+ + O_2 \rightarrow O_2^+ + O + 1.53 \text{ ,,} \quad (1b)$$
$$N^+ + O_2 \rightarrow NO^+ + O + 6.70 \text{ ,,} \quad (1c)$$
$$N^+ + O_2 \rightarrow NO + O^+ + 2.32 \text{ ,,} \quad (1d)$$
$$O_2^+ + N \rightarrow NO^+ + O + 4.23 \text{ ,,} \quad (1e)$$
$$O_2^+ + N_2 \rightarrow NO^+ + NO + 0.99 \text{ ,,} \quad (1f)$$
$$N_2^+ + O \rightarrow NO^+ + N + 3.10 \text{ ,,} \quad (1g)$$
$$N_2^+ + N \rightarrow N_2 + N^+ + 1.04 \text{ ,,} \quad (1h)$$
$$N_2^+ + O_2 \rightarrow NO^+ + NO + 4.50 \text{ ,,} \quad (1i)$$

The reaction $O_2^+ + N \rightarrow NO + O^+$ which has been considered by Hertzberg has been neglected in our analysis as the reaction is endothermic (Bates and Nicolet, 1961).

The following recombination reactions are considered :

$$O_2^+ + e \rightarrow O' + O' \quad \ldots \quad (2a)$$
$$N_2^+ + e \rightarrow N' + N' \quad \ldots \quad (2b)$$
$$NO^+ + e \rightarrow N' + O' \quad \ldots \quad (2c)$$

where (') signifies the excited species.

(Due to the high rate of photo-dissociation of negative ions, its presence has been neglected while considering the daytime equilibrium of positive ions).

The equations governing the daytime distribution of positive ions in the F-region are given by :

$$\dot{N} = J_1 a + J_2 b + J_3 c + J_4 d + J_5 e - N(\beta_1 x_1 + \beta_2 x_2 + \beta_3 x_5)$$
$$= q - N(\beta_1 x_1 + \beta_2 x_2 + \beta_3 x_5) \quad (3a)$$

where $n(O_2)$, $n(N_2)$, n(O), n(N) and n(NO) are denoted by a, b, c, d and e respectively. $n(O_2^+)$, $n(N_2^+)$, $n(O^+)$, $n(N^+)$, $n(NO^+)$ and n(e) are represented by x_1, x_2, x_3, x_4, x_5 and N respectively. J_1, J_2, J_3, J_4 and J_5 are the probabilities of photo-ionization of O_2, N_2, O, N and NO. β_1, β_2 and β_3 are the rate coefficients of dissociative recombination reactions (2a), (2b) and (2c) respectively.

If α_r ($r = 1, 2, 3, \ldots 9$) denotes the rate coefficients of ion-atom interchange reactions (1), we have

$$\dot{x}_1 = J_1 a + \alpha_2 a x_3 - x_1(\alpha_5 d + \alpha_6 b + \beta_1 N) \quad \ldots (3b)$$
$$\dot{x}_2 = J_2 b - x_2(\alpha_7 c + \alpha_8 d + \alpha_9 a + \beta_2 N) \quad \ldots (3c)$$
$$\dot{x}_3 = J_3 c + \alpha_4 a x_4 - x_3(\alpha_1 b + \alpha_2 a) \quad \ldots (3d)$$
$$\dot{x}_4 = J_4 d + \alpha_8 d x_2 - x_4(\alpha_3 a + \alpha_4 a) \quad \ldots (3e)$$
$$\dot{x}_5 = J_5 e + \alpha_1 b x_3 + \alpha_3 a x_4 + x_1(\alpha_5 d + \alpha_6 b) + x_2(\alpha_7 c + \alpha_9 a) - \beta_3 N x_5 \quad \ldots (3f)$$

Equations (3b) to (3f) can be written as,

$$\dot{x}_1 = -m_{11} x_1 + m_{13} x_3 + l_1 \quad \ldots (4a)$$
$$\dot{x}_2 = -m_{22} x_2 + l_2 \quad \ldots (4b)$$
$$\dot{x}_3 = -m_{33} x_3 + m_{34} x_4 + l_3 \quad \ldots (4c)$$
$$\dot{x}_4 = -m_{44} x_4 + m_{42} x_2 + l_4 \quad \ldots (4d)$$
$$\dot{x}_5 = -m_{55} x_5 + m_{51} x_1 + m_{52} x_2 + m_{53} x_3 + m_{54} x_4 + l_5 \quad \ldots (4e)$$

where,

$$m_{11} = \alpha_5 d + \alpha_6 b + \beta_1 N$$
$$m_{22} = \alpha_7 c + \alpha_8 d + \alpha_9 a + \beta_2 N$$
$$m_{33} = \alpha_1 b + \alpha_2 a$$
$$m_{44} = \alpha_3 a + \alpha_4 a \qquad m_{53} = \alpha_1 b$$
$$m_{55} = \beta_3 N \qquad m_{54} = \alpha_3 a$$
$$m_{13} = \alpha_2 a \qquad l_1 = J_1 a$$
$$m_{34} = \alpha_4 a \qquad l_2 = J_2 b$$
$$m_{42} = \alpha_8 d \qquad l_3 = J_3 c$$
$$m_{51} = \alpha_5 d + \alpha_6 b \qquad l_4 = J_4 d$$
$$m_{52} = \alpha_7 c + \alpha_9 a \qquad l_5 = J_5 e.$$

Denoting the initial concentrations of positive ions x^0_i ($i = 1, 2, 3, 4, 5$), the solutions of equation (4) are obtained. For example, the solution of equation (4*b*) is given by

$$x_2 = \frac{l_2}{m_{22}} - \left(\frac{l_2}{m_{22}} - x_2^0\right) e^{-m_{22}t} \qquad \ldots (5)$$

In a similar manner the other solutions can be obtained.

Substituting the numerical values of various quantities in equation (5), daytime distribution of ions at different altitudes can be calculated.

CALCULATION OF ION DISTRIBUTION FOR THE F-REGION

Using equation (5), the ion distribution in the F-region is calculated after determining the rates of production of ions and constants (m^s) at different altitudes.

(a) *Rates of production of ions*

To determine the rates of production of ions, the following procedure is adopted :

The number of photons, $n(h\nu)_z$, incident per sq. cm per second corresponding to frequency ν at an altitude z is given by,

$$n(h\nu)_z = n(h\nu)_\infty \exp\left(-\sum_i n(i)_z H_{iz}\sigma_{\nu i}\right) \qquad \ldots (6)$$

where,

$n(h\nu)_\infty$—number of photons of frequency ν at the top of the atmosphere,

$n(i)_z$—particle concentration of i-th constituent at the altitude z,

H_{iz}—scale height of the i-th constituent at the altitude z,

and $\sigma_{\nu i}$—absorption cross-section corresponding to frequency ν for the i-th constituent.

The wavelength region 1000—100 Å, which is mainly responsible for the photo-ionization of F-region, is divided into intervals such that within each interval the absorption coefficient is practically constant. For oxygen and nitrogen atoms, the photo-ionization cross-sections given by Dalgarno and Parkinson (1960) were utilized. The absorption cross-section of O_2 upto 500A as given by Watanabe (1958) and below it those obtained by Weissler *et al* (1955), are taken. For N_2 upto 303A absorption cross-section given by Watanabe (1958) and below it those obtained by Weissler *et al* (1952) are assumed. Miller's (1957) model of atmospheric composition and scale heights as given in ARDC model atmosphere 1959 are used. Thorough mixing is assumed upto 150 km for O_2 and upto 180 km for N_2, N and O. Above these heights diffusive equilibrium is considered. Values of $n(h\nu)$ have been taken from the results of rocket-borne experiments extrapolated by Hinteregger (1961).

The rates of production of photo-ions produced due to ionization of the i-th constituent is calculated from the expression,

$$\sum_{\nu} n(h\nu)_z \cdot n(i)_z \cdot \sigma_{\nu i} = n(i)_z \Sigma_{\nu} n(h\nu)_z \cdot \sigma_{\nu i} = n(i)_z \cdot J_{iz}$$

where, J_{iz} is the probability of ionization of the i-th constituent at the altitude z.

The absorption coefficients and atmospheric composition as mentioned above have been used. The calculated rates of production of ions are given in Table I. The production of NO^+ ions due to photo-ionization above 120 km has been neglected owing to the small density of NO molecules. The ionization due to solar X-rays is significant only below 160 km and is shown in Table I.

(b) *Determination of constants* m^s

The constants m^s are controlled by rate coefficients of ion-atom interchange (α^s) and dissociative recombination (β^s) processes. The electron density used in the calculations has been taken from rocket-borne experimental data obtained by Nisbet (1960). It is assumed that the rate coefficients of ion-atom interchange and dissociative recombination processes are constant. Taking known values of β^s, α^s are fixed at 150 km so as to give the observed percentage distribution of ions (Johnson, 1961). Using above values of α^s and β^s, equilibrium time (Table II) is calculated for each ion and it is found that except for O^+ and NO^+ above 320 km, equilibrium is established within a short time. Taking above values of α^s and β^s, the ion distributions are calculated up to 300 km. These distributions are compared with those obtained from Johnson's data (Fig. 1). It is found that,

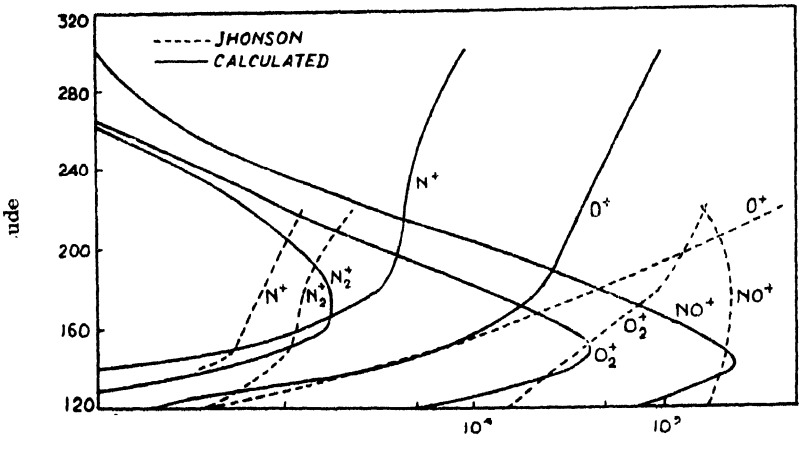

Fig. 1. The altitude variation of positive ions concentration in the F-region assuming the same rate coefficients for every altitude as that fixed for 150 km.

(i) the calculated distributions do not agree with the observed distributions below and above 150 km and that the discrepancy between two distributions is greater above 180 km, and

TABLE I

Altitude (km)	Rate of production of ions (cm^3 sec^{-1}) by Solar ultraviolet radiations				Rate of production of ions (cm^3 sec^{-1}) by Solar X-rays			
	O_2^+	N_2^+	O^+	N^+	O_2^+	N_2^+	O^+	N^+
1200	1.28×10^{2}	1.06×10^{2}	1.66×10^{1}	1.95×10^{-1}	1.62×10^{2}	5.53×10^{2}	8.45×10^{2}	2.38
140	1.78×10^{2}	5.60×10^{3}	1.05×10^{3}	6.02×10^{1}	4.12×10^{1}	1.68×10^{2}	2.45×10^{2}	2.95
150	1.40×10^{2}	5.94×10^{3}	1.06×10^{3}	9.15×10^{1}	1.49×10^{1}	7.70×10^{1}	1.12×10^{2}	1.90
160	6.34×10^{1}	4.21×10^{3}	6.44×10^{2}	8.73×10^{1}	6.09	3.77×10^{1}	5.53×10^{1}	1.34
180	2.86×10^{1}	2.09×10^{3}	2.52×10^{2}	5.94×10^{1}				
200	1.08×10^{1}	8.99×10^{2}	2.16×10^{2}	3.88×10^{1}				
220	4.20	3.95×10^{2}	1.32×10^{2}	2.48×10^{1}				
240	1.77	1.83×10^{2}	8.21×10^{1}	1.62×10^{1}				
260	8.01×10^{-1}	9.09×10^{1}	5.27×10^{1}	1.09×10^{1}				
280	3.87×10^{-1}	4.74×10^{1}	3.52×10^{1}	7.54				
300	1.98×10^{-1}	2.61×10^{1}	2.43×10^{1}	5.36				
320	1.06×10^{-1}	1.51×10^{1}	1.72×10^{1}	3.94				
340	5.95×10^{-2}	9.00	1.24×10^{1}	2.94				
360	3.46×10^{-2}	5.54	9.13	2.24				
380	2.07×10^{-2}	3.51	6.86	1.72				
400	1.29×10^{-2}	2.30	5.26	1.35				

TABLE II

Equilibrium time (the time required to attain 1/e-th of the equilibrium concentration) for various ions

Altitude (km)	Time of equi ibrium for O_2^+	N_2^+	O^+	N^+	NO^+
120	22.4 sec	0.22 sec	0.65 sec	11.96 sec	0.65 sec
160	1.03 mts	0.56 sec	28.8 sec	1.55 sec	28.8 sec
200	2.81 „	1.02 „	5.4 mts	9.35 „	5.4 mts
240	1.99 „	1.61 „	35.0 „	1.03 mts	25.0 „
280	1.15 „	1.43 „	2.65 hrs	4.80 „	2.65 „
320	1.16 „	1.57 „	9.53 „	17.67 „	9.53 „
360	1.61 „	2.25 „	28.61 „	54.33 „	28.61 „
400	2.21 „	3.12 „	75.56 „	2.44 hrs	75.56 „

TABLE III

Altitude Variation of the Effective Recombination Coefficient α_{eff} (assuming $\beta_1^* = 1\times10^{-8}$cm^3/sec, $\beta_2^* = 4\times10^{-7}$ cm^3/sec and β_3^* as follows)

Altitude (km)	β_3(cm^3 sec^{-1})	α_{eff}(cm^3 sec^{-1}) Calculated	Havens *et al* (1954)
120	5.5×10^{-8}	5.0×10^{-8}	1.6×10^{-8}
140	$1.5=10^{-7}$	1.4×10^{-7}	—
150	1.3×10^{-7}	1.1×10^{-7}	1.6×10^{-8}
200	7.5×10^{-9}	4.4×10^{-9}	2.0×10^{-9}
250	3.0×10^{-9}	1.1×10^{-10}	1.3×10^{-10}
300	—	1.1×10^{-11}	1.6×10^{-11}
340	—	7.2×10^{-12}	5.0×10^{-12}
360	—	7.5×10^{-12}	3.0×10^{-12}
400	—	7.4×10^{-12}	1.8×10^{-12}

*Bortner and Baulknight (1961) have recently reported that β_1 and β_2 are temperature dependent whereas β_3 is independent of temperature. Their values are $\beta_1 = 9.1\times10^{-5}\ T^{-1}$, $\beta_2 = 1.1\times10^{-5}\ T^{-1/2}$ and $\beta_3 = 1\times10^{-8}$.

(ii) total number of positive ions is not equal to electron density.

From above, it is clear that m^s should vary with altitude. It is generally agreed that dissociative recombination coefficients are constant (their dependance on temperature is not known, Bates and Nicolet, 1960). Therefore, keeping β^s constant, α^s are fixed at each altitude to reproduce the observed ion distribution. It is found that α^s increase upto 180 km and then decrease. β^s can be kept constant only upto 180 km and above it they have to be decreased together with α^s. It should, however, be noted that effective recombination coefficient α_{eff} calculated from the expression,

$$\alpha_{eff} = \frac{\beta_1 n(O_2^+) + \beta_2 n(N_2^+) + \beta_3 n(NO^+)}{N}$$

agrees with those obtained by Havens *et al* (1954) except for the F_1-layer (Table III)

The rate coefficient for dissociative recombination of NO^+, β_3, could not be kept constant for all altitudes as assumed earlier. From Eqn. (3), it is seen that β_3 appears in the last equation and is therefore solved at the end. If a fixed value of β_3(say 1×10^{-8} cm^3/sec) is taken, the density of NO^+ ions at lower altitudes exceeds the electron density which is not tenable. β_3 has to be taken as high as 1.5×10^{-7} cm^3/sec at 140 km which is decreased to 2.3×10^{-8} cm^3/sec at 180 km in order to reproduce the observed distribution of NO^+. β_3 as low as 3.0×10^{-9} cm^3/sec, as suggested by Bates and Nicolet (1960), reproduces the observed distribution of NO^+ above 180 km and is kept constant for higher altitudes.

Burkard (1962) has suggested that β_3 should vary approximately linearly with temperature. In our analysis, we obtained that β_3 increases approximately linearly with temperature (upto about 140 km) and that for higher temperatures it decreases exponentially with temperature. This variation in β_3 may be due to the fact that we have assumed β_1 and β_2 independent of temperature whereas Bortner and Baulknight (1961) have shown that $\beta_1 = 9.1\times10^{-5}\,T^{-1}$, $\beta_2 = 1.1\times10^{-5}T^{-1/2}$ and $\beta_3 = 1.0\times10^{-8}$ cm^3/sec. The values of β_1 and β_2 used by us are, of course, of the same magnitude (in the F-region) as those reported by Bortner and Baulknight. However, further investigations are needed to verify these results.

From above, it seems that the situation above 180 km is not so simple as assumed above. It has been shown by many workers that above 180 km, diffusion and ion drift play an important role in the distribution of ions and electrons.

RATE COEFFICIENTS OF ION—ATOM INTERCHANGE REACTIONS

The rate coefficients of ion-atom interchange reactions except for α_2 are not determined experimentally. α_2 has been measured by Dickinson and Sayers (1960) to be $2.5 \perp 0.4 \times 10^{-11}$ cm^3/sec at 300°K. Since the ion atom interchange is a two-body process, its coefficient is likely to increase with temperature analogous to the rate coefficient of two-body chemical process. Bates and Nicolet (1960) pointed out that the rate coefficients of these reactions should be an increasing function of temperature and that these might possess some activation energy or steric hinderance. We have, therefore, assumed that the rate coefficients vary as,

$$\alpha_r = C_r \exp(-E_r/RT)$$

where, $r = 1, 2, 3, \ldots 9$, R is the gas constant and E_r represents the activation energy. Graphs between $\log_e \alpha_r$ and $1/T$ (the values of α_s are already fixed between 120—180 km) are found to be straight lines (Fig. 2). They are then extrapolated to obtain α^s at higher altitudes (Table IV). From Fig. 2, it is seen that the magni-

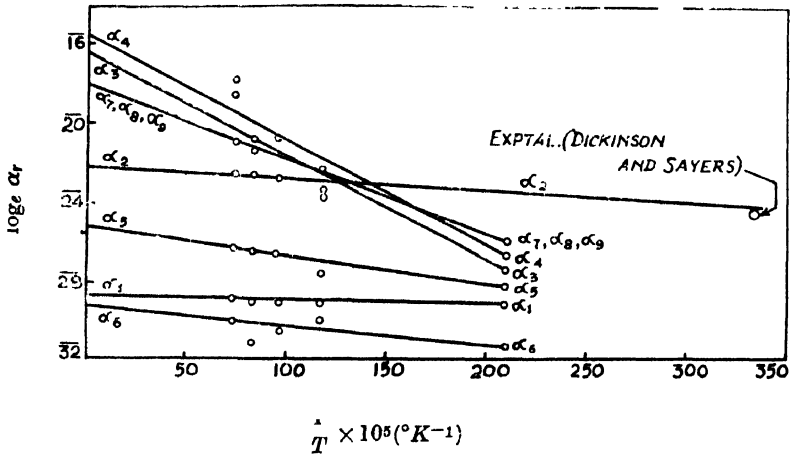

Fig. 2. The variation of rate coefficients of ion-atom interchange reactions with temperature.

tudes of α^s do not differ appreciably above 200 km as $1/T$ remains approximately constant (ARDC Model Atmosphere, 1959). The values of α^s at 200 km have been used for higher altitudes.

From Table IV, it is seen that the rate coefficients α_3 and α_4 of ion-atom interchange reaction involving N^+ ion and O_2 vary between 10^{-12} and 10^{-9} cm^3/sec and that their activation energies are very high i.e. 7.74 Kcal. Since the densiyt of N^+ ions is very small, the effect of these rate coefficients on the distribution of ions is small. The values of α_3 and α_4 above 180 km agree with the theoretically calculated order of 10^{-9} cm^{-3}/sec for such a reaction (Bates and Nicolet, 1960). However, such a high value of rate coefficients has not been obtained for other

TABLE IV

Temperature variation of rate coefficients of ion-atom interchange reactions. Quantities in paranthesis represent activation energies of reactions in Kcal

Altitude (km)	Temperature (°K)	α_1* (0.47)	α_2* (0.95)	α_3 (7.74)	α_4 (7.73)	α_5 (1.59)	α_6* (1.61)	α_7*,α_8,α_9* (3.56)
1200	477.0	2.0×10^{-13}	7.0×10^{-11}	1.26×10^{-12}	2.80×10^{12}	1.0×10^{13}	9.0×10^{-14}	1.5×10^{-11}
140	849.9	2.3×10^{-13}	1.2×10^{-10}	1.61×10^{-10}	3.31×10^{10}	5.0×10^{12}	7.0×10^{-13}	2.0×10^{-10}
150	1031.0	2.4×10^{-10}	1.3×10^{-10}	5.07×10^{-10}	1.02×10^{-9}	6.0×10^{-12}	9.0×10^{-13}	4.0×10^{-10}
160	1207.0	2.5×10^{-13}	1.5×10^{-10}	1.0×10^{-9}	2.00×10^{-9}	6.5×10^{-12}	9.5×10^{-13}	5.5×10^{-10}
180	1371.0	3.0×10^{-13}	1.6×10^{-10}	1.77×10^{-9}	3.76×10^{-9}	7.0×10^{-12}	1.0×10^{-12}	7.5×10^{-10}
200	1404.0	3.0×10^{-13}	1.6×10^{-10}	1.77×10^{-9}	3.76×10^{-9}	7.0×10^{-12}	1.0×10^{-12}	7.5×10^{-10}

*M. H. Bortner and C. W. Baulknight (1961) in the Scientific Report No. 1 'Deionization kinetics', Contract No. AF19(604)—8820, AF Cambridge Research Center, Bedford, Mass., U.S.A. have reported that $\alpha_1 = 5.0\times10^{-12}$, $\alpha_2 = 2.5\times10^{-11}$, $\alpha_6 = 0$, $\alpha_7 = 1.0\times10^{-11}$ and $\alpha_9 = 0$ cm^3/sec.

reactions which usually vary between 10^{-13} to 10^{-10} cm^3/sec. Table IV further shows that reactions involving the same ion have nearly equal activation energies.

The value of α_2 is found to be 3.07×10^{-11} cm^3/sec at 300°K which agrees with the experimentally determined value ($2.5 \pm 0.4\times10^{-11}$ cm^3/sec) of Dickinson and Sayers (1960) at the same temperature. We obtained the values of α_1 and α_2 for 1415°K to be 3.0×10^{-13} cm^3/sec and 1.6×10^{-10} cm^3/sec respectively. Bates and Nicolet (1960) also found that α_1 and α_2 differ by two orders at 250 km which, according to ARDC model atmosphere, 1959, has a temperature of 1415°K. It should, however, be noted that the reactions (1*a*) and (1*b*) prossess very small activation energies, and therefore, they seem to be important reactions in the ionosphere. Reaction (1a) is probably very important in the F_2 region where O^+ ions and N_2 molecules are in abundance. Through this reaction, O^+ ions are converted into NO^+ ions which readily dissociates ($\beta_3\sim1.0\times10^{-8}$ cm^3/sec) into nitrogen and oxygen atoms. The intensity variation of λ 6300 with altitude in twilight and day airglow may be able to show the importance of the two controlling processes (1a) and (1b) in the higher region of the F-layer.

EFFECT OF DIFFUSION ON ION DISTRIBUTIONS*

Using the above values of α^s and β^s ion densities are calculated, which are found to be lower than the observed values above 180 km. The discrepancy can be accounted for, if diffusion and vertical drift are taken into consideration.

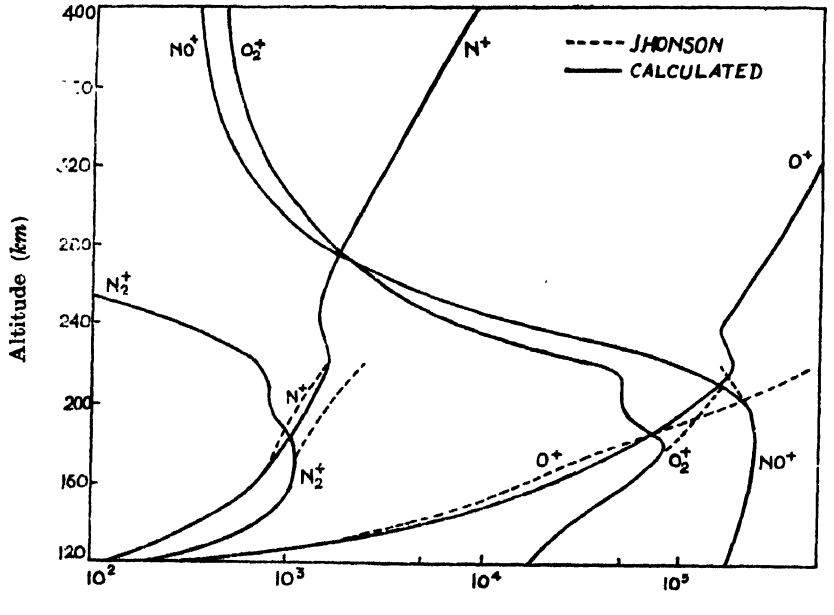

Fig. 3. The distribution of positive ions in the F-region of the ionosphere.

*The diffusion of ions and electrons has been considered qualitatively by K. D. Sharma (D. Phil. thesis, University of Allahabad, 1962).

It is found that for altitudes between 200 and 220 km (above which the observed percentage ion distribution is not available), the discrepancy can be partially explained if the contribution due to diffusion is taken in the same proportion as the percentage of ions (Johnson, 1961) Fig. 3. To obtain more accurate agreement the vertical drift should also be considered.

EFFECT OF THE VERTICAL DRIFT

From the expression for the vertical drift, (Ferraro, 1961), we have

$$W = \frac{\delta}{\delta z} \quad (Nw).$$

If it is assumed that the drift velocity, ω, is constant at every altitude, we then have,

$$W = \frac{\delta N}{\delta z}$$

By assuming that the difference in the total number of positive ions and the electron density at 200 km is caused by the vertical drift, ω is found to be about 16 metres/sec. This is of the same order as obtained by Ferraro (1961) in the F-region (10 metres/sec).

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